Along-strike variations of P-T conditions in accretionary wedges and syn-orogenic extension, the HP-LT Phyllite-Quartzite Nappe in Crete and the Peloponnese

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Abstract

Syn-orogenic detachments in accretionary wedges make the exhumation of high-pressure and low-temperature metamorphic rocks possible with little erosion. The velocity of exhumation within the subduction channel or the accretionary complex, and thus the shape of *P–T* paths, depend upon the kinematic boundary conditions. A component of slab retreat tends to open the channel and facilitates the exhumation. We document the effect of slab retreat on the shape of *P–T* paths using the example of the Phyllite–Quartzite Nappe that has been exhumed below the Cretan syn-orogenic detachment during the Miocene in Crete and the Peloponnese. Data show a clear tendency toward colder conditions at peak pressure and during exhumation where the intensity of slab retreat is larger. This spatial evolution of *P–T* gradient is accompanied with an evolution from a partly coaxial regime below the Peloponnese section of the detachment toward a clearly non-coaxial regime in Crete.

Keywords : High pressure and low temperature metamorphic rocks; Exhumation; Slab retreat; Detachment; Crete; Peloponnese

Exhumation of blueschists and eclogites is a complex process that involves a succession of mechanisms from the depth of the subduction channel to the surface. Tectonic processes within the subduction channel or the accretionary wedge compete with erosion ([Platt, 1986], [Cloos and Shreve, 1988], [Platt, 1993], [Avigad et al., 1997], [Brandon et al., 1998], [Ernst and Liou, 2000], [Jolivet et al., 2003] and [Ring, and Layer, 2003]). In the case of the Aegean region (Fig. 1), erosion played a minor role in the Tertiary, this is attested by the ubiquitous presence on top of the high-pressure and low-temperature HP–LT units of the Upper Cycladic Nappe that does not show any high pressure metamorphic overprint (Jolivet et al., 2003). In the Aegean the most spectacular structures related to exhumation were formed during post-orogenic extension ([Lister et al., 1984], [Gautier et al., 1993] and [Gautier and Brun, 1994]) and the formation of extensional metamorphic domes. In this case the HP–LT parageneses are overprinted by high temperature ones. A significant example is the Naxos dome where remains of blueschists and eclogites are found only in the periphery of the dome (Avigad, 1998) whereas the core shows amphibolites and migmatites ([Feenstra, 1985], [Vanderhaeghe, 2004] and [Duchêne et al., 2006]). In the example of the Tinos core complex, although less severe, the evolution toward high temperatures is clearly seen in the late stages in the *P*–*T*–time path (Parra et al., 2002b). In such cases extracting from the geological record the part of the tectonic history due solely to syn-orogenic exhumation is not always straightforward, even more when the transition from syn-orogenic to post-orogenic exhumation occurred in a short time span not easily deciphered with radiochronological studies. A very significant part of the exhumation was however achieved within the subduction zone before the formation of the Aegean Sea ([Avigad et al., 1997], [Trotet et al., 2001a], [Jolivet et al., 2003], [Ring, and Layer, 2003] and [Jolivet and Brun, 2008]).

Blueschists and eclogite massifs of the Mediterranean region show a wide range of kinematic and *P*–*T* evolution that are due to different geodynamic contexts (Jolivet et al., 2003). Kinematic boundary conditions, whether the slab is retreating or not for instance, are of paramount importance for the dynamics of the accretionary complex as shown by numerical models and regional syntheses (Beaumont et al., 1999) besides other major factors such as the lithologic nature of the subducted material (Goffé et al., 2003).

The external Hellenic arc was formed during the rifting of the Aegean Sea. The HP–LT parageneses of the Phyllite–Quartzite Nappe are contemporaneous with the HT–LP parageneses of the Cyclades and exhumation is associated with the activity of a large-scale detachment. Little post-orogenic extension is recorded there and we can thus use this example to study the dynamics of syn-orogenic exhumation. The Phyllite–Quartzite Nappe (Fig. 1) offers the opportunity to study along strike variations of this dynamics with a single, mostly pelitic protolith, thus eliminating possible lithological effects.

1. Geodynamic setting

The Phyllite–Quartzite Nappe (PQ) is an external tectonic unit of the Hellenides (Fig. 1 and Fig. 2). It belongs to the Hellenic nappe stack that is recognized from continental Greece to Crete and the Cyclades ([Bonneau, 1982], [Seidel et al., 1982] and [Bonneau, 1984]). It has recorded an Oligo-Miocene high pressure and low temperature (HP–LT) metamorphic stage contemporaneous with a high temperature and low pressure (HT–LP) event in the Cyclades further north. An earlier episode of nappe stacking in the Eocene during the subduction of the northern margin of the Apulian continental block and the Pindos ocean below the southern margin of Eurasia led to the formation of the Hellenides and the burial and exhumation of a first HP–LT nappe, the Cycladic Blueschists (Fig. 1 and Fig. 2) ([Blake et al., 1981], [Bonneau and Kienast, 1982], [Jolivet et al., 2003], [Ring, and Layer, 2003] and [Jolivet and Brun, 2008]). From 30 to 35 Ma onward the subduction regime changed and the Aegean Sea started to rift in the backarc region of the Hellenic subduction, while the front of subduction migrated continuously southward, following the retreat of the African slab (Jolivet and Faccenna, 2000). Eocene HP–LT parageneses were then reworked by HT–LP ones in the Cyclades while the southern portions of the Apulian block were dragged in the subduction zone and subjected to a new episode of blueschists-facies metamorphism.

The Phyllite–Quartzite (PQ) Nappe can be observed all along the external Hellenic arc from Crete to the Peloponnese. Although the structural position of the PQ Nappe within the orogenic wedge is similar along strike, the tectonic environment evolves gradually from east to west. Crete is located in the centre of the arc along a N–S transect that has recorded a maximum of slab retreat and extension. North of Crete the Cretan Sea is widely open and further north the metamorphic core complexes of Naxos and Mykonos show deep portions of the Aegean crust exhumed during the Oligo-Miocene. A section perpendicular to the arc passing through the northern Peloponnese shows much less post-orogenic extension, the Cretan Sea is absent there and the metamorphic core complexes of Evia and Andros have exhumed only greenschist-facies metamorphic rocks during the Oligo-Miocene. The finite amount of slab retreat is smaller there than in the centre of the Hellenic arc.

Slab retreat started, or accelerated in the Late Eocene–Early Oligocene and it is still active now. The subduction and exhumation of the PQ Nappe were thus contemporaneous of slab retreat thus making it a good candidate to study the influence of the velocity of slab retreat on P–T conditions in the orogenic wedge.

2. The Phyllite–Quartzite Nappe

The PQ Nappe (Fig. 3 and Fig. 4) ([Creutzburg, 1977] and [Bonneau, 1984]) is sandwiched between the overlying Gavrovo–Tripolitza Nappe (GT) made of thick platform carbonates ranging in age from Triassic to Eocene overlying low-grade metapelites and volcanic rocks of Triassic age (Tyros beds), and the underlying Plattenkalk Nappe (PK), also referred to as the Taygetos group (Deckert et al., 1999), a thick accumulation of Triassic to Eocene platform and pelagic limestones, more external than the GT Nappe in the paleogeography of the Hellenides ([Krahl et al., 1983], [Thiébault and Triboulet, 1983], [Deckert et al., 1999] and [Robertson, 2006]). The thickness of the Tyros beds, also named the Ravdoucha beds (Seidel et al., 2005), is highly variable, a few hundred meters are usually recognized. The PQ Nappe also shows very large thickness variations. In Crete it can be 1 km thick in the west and it can be sometimes totally missing between the Plattenkalk and the GT Nappe. The PQ Nappe itself is made of Triassic metapelites, conglomerates, quartzites and minor limestones, including slices of paleozoic basement found both in Crete and the island of Kithira ([Romano et al., 2004] and [Seidel et al., 2006]). The average resistance of the PQ Nappe was probably much smaller than that of the PK and GT Nappes below and above (Stöckhert et al., 1999) and it consequently localized a large part of the deformation during burial and exhumation. It is recognized in the Zaroukla–Feneos window, in the Parnon and Taygetos massifs in the southern Peloponnese, in the island of Kithira and in Crete. A high-pressure and low-temperature (HP–LT) metamorphic imprint dated to the late Oligocene–Early Miocene has been recognized in Crete and the southern Peloponnese ([Seidel et al., 1982], [Papanikolaou and Skarpelis, 1986], [Theye and Seidel, 1991], [Theye et al., 1992], [Theye and Seidel, 1993], [Jolivet et al., 1996], [Thomson et al., 1998], [Trotet, 2000] and [Zulauf et al., 2002]). Conversely, the GT Nappe never shows any evidence of high pressure recrystallization. This observation and the presence of a distinct shear zone at the top of the PQ Nappe in Crete led [Jolivet et al., 1994b], [Jolivet et al., 1996], [Kilias et al., 1994] and [Fassoulas et al., 1994] to propose the existence of a major north-dipping extensional detachment between the GT and PQ Nappes. The formation of Miocene E–W basins is related to the development of this detachment ([Ring et al., 2001a], [van Hinsbergen and Meulenkamp, 2006] and [Seidel et al., 2007]). A slightly different model involving more compression but with a normal fault system at the top was recently proposed by Chatzaras et al. (2006).

3. Tectonic and metamorphic evolution of the Phyllite–Quartzite Nappe

The rather homogeneous metapelite-rich lithology of the PQ Nappe allows a precise description of its metamorphic evolution (Fig. 5 and Fig. 6). After the first finding of Fe–Mg carpholite in the PQ Nappe as well as aragonite and Fe–Mg carpholite in the PK Nappe in Crete ([Seidel, 1978] and [Seidel et al., 1982]), HP–LT metamorphic conditions have been described in the whole PQ Nappe from the Peloponnese to Crete and *P–T–t* paths were estimated ([Thiébault and Triboulet, 1983], [Theye and Seidel, 1991], [Theye et al., 1992], [Theye and Seidel, 1993], [Bassias and Triboulet, 1994], [Jolivet et al., 1996] and [Trotet et al., 2006]). The maximum pressure and temperature conditions are found in western Crete (16–18 kbar and 400 °C) and the southern Peloponnese (16–18 kbar and 500–550 °C) with a significantly warmer evolution in the Peloponnese ([Jolivet et al., 1996] and [Trotet et al., 2006]). Pressure then decreases eastward in Crete and northward in the Peloponnese.

3.1. Crete

The absence of any HP–LT parageneses in the Gavrovo–Tripolitza Nappe of Crete, including the Tyros Beds, is evidence for a major tectonic gap across the basal thrust contact. The main contact is located between a lower unit where metapelites contain abundant Fe–Mg carpholite and chloritoid associated to phengites (the Phyllite–Quartzite Nappe proper) and the base of the Gavrovo–Tripolitza Nappe where similar metapelites (the Tyros Beds) do not show any significant metamorphic recrystallization. The absence of Fe–Mg carpholite in the Tyros beds indicates a maximum pressure below ~ 6 kbar while the parageneses below the contact yield pressure estimates of up to 16 kbar (Jolivet et al., 1996). The absence of Fe–Mg carpholite is in line with the apatite fission-track thermochronological history of the GT Nappe showing that this unit has remained in the upper 4 km of the crust since ~ 35 Ma (Thomson et al., 1998). A minimum pressure gap of ~ 10 kbar suggests that ~ 30 km or more are missing. This has been used to document the existence of a major detachment (called the Cretan detachment hereafter) ([Fassoulas et al., 1994], [Jolivet et al., 1994b], [Jolivet et al., 1996] and [Ring et al., 2001b]). The good preservation of Fe–Mg carpholite close to the detachment shows that the rocks were exhumed quickly after their peak of pressure. Kinematic indicators show a clear top-to-the-north shear sense along the detachment (Fig. 2). The spatial variation of the shape of *P–T* paths in the PQ Nappe furthermore shows that, close to of the GT Nappe, the *P–T* evolution is colder than further down, with a clear cooling during decompression and the preservation of Fe–Mg carpholite whereas deeper units show an almost complete replacement of carpholite by chloritoid during an isothermal decompression (Jolivet et al., 1996). This observation is in favour of the extensional nature of the GT/PQ contact, with a cooling of the lower unit near its contact with the upper colder unit (see also Jolivet et al., 1998). The downward temperature increase is recorded only in western Crete. Eastern Crete outcrops do not show this evolution and the peak of pressure is associated to a lower temperature ([Jolivet et al., 1996] and [Zulauf et al., 2002]). This is in line with the fission-track records of zircons that show young ages (~ 17 Ma) only in the west where higher temperature has completely reset fission track ages, in eastern Crete where the temperature of metamorphism is lower zircon FT ages range from 414 to 145 Ma (Brix et al., 2002).

A more recent analysis of the distribution of temperatures across the Cretan nappe stack showed a limited temperature difference across the Cretan detachment and the authors concluded that this structure has not contributed to more than 7 km of exhumation (Rahl et al., 2005). However, the proxy used in Rahl et al. (2005) is *temperature* whereas the *maximum pressure* should be used to estimate the amount of relative vertical displacement. It is indeed very unlikely that the temperature gradient was continuous across the nappe stack and thus a temperature difference cannot be used to estimate the amount of exhumation across a tectonic contact. We shall thus consider in the following that the Cretan detachment was responsible for most of the exhumation of the PQ Nappe.

$^{40}Ar/^{39}Ar$ dating of phengites suggest that the peak of pressure was attained some 24 Ma ago and that the bulk of the PQ Nappe was exhumed in the upper crust some 15 Ma ago (Jolivet et al., 1996). The detachment was thus active in the early Miocene. Step-heating $^{40}Ar/^{39}Ar$ ages on single grains of phengites in Crete provide a time range between 25 and 15 Ma for the whole history of exhumation of the PQ Nappe before entering the brittle field (Jolivet et al., 1996). Thermochronological data using fission-tracks suggest that most of the exhumation was accomplished before 19 Ma that is shortly after the peak of pressure, that the rate of exhumation was high (≥ 4 km/Myr) and that 85–90% of the exhumation was achieved by tectonic processes instead of by erosion ([Thomson et al., 1998] and [Brix et al., 2002]). This indicates that most of the normal sense relative displacement along the detachment occurred before whole crust extension.

3.2. The southern Peloponnese

The same detachment can be recognized in the Peloponnese with a smaller pressure gap than in Crete (Fig. 3 and Fig. 4) (NB, n order to emphasize the lateral continuity of structures from Crete to the Peloponnese we use the same name, “Cretan Detachment”, in both regions). As in Crete the absence of HP–LT parageneses in the overlying GT Nappe and the ubiquitous presence of Fe–Mg carpholite, chloritoid, glaucophane and locally garnet (Fig. 5) in the PQ Nappe suggests the presence of detachments. The distribution of temperature and the shapes of *P*–*T* paths in the PQ Nappe in the footwall of the Cretan Detachment are however more complex with high pressure units resting on top of lower pressure ones (Fig. 4). It thus seems that a more complex post-metamorphic imbrication of tectonic units occurred below the detachment within the accretionary complex than in Crete. Based on the petrological study of Trotet et al. (2006) we present here a set of new results concerning the deformation and *in-situ* $^{40}Ar–^{39}Ar$ ages of the PQ Nappe in the southern Peloponnese. Several studies have attempted to classify the various lithologies of the PQ Nappe into coherent units ([Deckert et al., 1999], [Papanikolaou and Skarpelis, 1986] and [Aleweld et al., 1994]). Part of the so-called PQ Nappe is also now considered as the basement of the PK Nappe ([Dittmar and Kowalczyk, 1991] and [Deckert et al., 1999]). One of the most striking characteristics of these different units is their contrasting metamorphic evolution. We thus think that they evolved differently for a part of their history and that they should be considered as separate tectonic units. We therefore use in the whole paper the unit names and the lithological description of Trotet et al. (2006) as it is based on the lithology of the protoliths and their metamorphic evolution.

In the Taygetos and Parnon massifs Trotet et al. (2006) recognized four units within the PQ Nappe with different metamorphic histories (Fig. 4), (1) blueschist-facies metaconglomerates and micaschists with Fe–Mg carpholite and chloritoid hereafter named the *Metaconglomerate Unit*, (2) blueschist-facies micaschists with chloritoid and lenses of glaucophanites, hereafter named the *Blueschist Unit*, (3) greenschist-facies micaschists with chloritoid and albite (*Alagonia Unit*). In the Taygetos one more unit comes near the top of the stack below the GT Nappe, it contains (4) HT blueschist-facies micaschists with glaucophane and garnet (*Lada Unit*). The abrupt transitions between units suggest that the present stack is post-metamorphic. The basal contact of the PQ Nappe with the PK appears to cut across the tectonostratigraphy of the PQ Nappe and is associated with low-pressure metamorphic recrystallisation. The absence of characteristic HP–LT parageneses in the upper part of the underlying Ionian (PK) nappe further suggests that the basal contact of the PQ Nappe is a post-metamorphic thrust and that the PK Nappe was buried to shallower depth in the Peloponnese than in Crete. Blumör et al. (1994) have however described the presence of Fe–Mg carpholite in the Kastania Phyllite (not to be confused with the sample from Kastania, located east of the Zaroukla–Feneos window, see below) which they attribute to the base of the PK unit, leading to *P*–*T* estimates around 7–8 kbar and 310–360 °C.

The analysis of retrograde ductile deformation in the PQ Nappe shows a ubiquitous E–W stretching lineation and bivergent kinematic indicators indicating conjugate shear zones (Fig. 3). The finite fabric, at the scale of the southern Peloponnese, is more symmetrical than in Crete without a clear predominance of one shear sense at the scale of the southern Peloponnese. It should be noticed however that the E–W direction observed at present must be rotated to restore the Neogene 50° clockwise rotation ([van Hinsbergen and Meulenkamp, 2005] and [Kissel and Laj, 1988]) of the Peloponnese and the actual stretching direction was thus NE–SW.

The stretching lineation is associated with blueschist and greenschist parageneses and the shear sense seems constant during decompression. There thus seems to be a continuum in the direction of shear from the peak pressure throughout the exhumation process. Kinematic indicators are top-to-the-west west of Mt Taygetos and top-to-the-east elsewhere. The presence of the greenschist Alagonia unit at the top of the PQ stack shows a smaller pressure gap than in Crete. The contact with the GT Nappe has thus accommodated less displacement in the Peloponnese than in Crete and the deformation is distributed across a greater thickness. The PQ Nappe as a whole is thus sandwiched between a late thrust at its base and a subtractive contact at the top.

P–*T* estimates and the shape of *P*–*T* paths (Fig. 6) have been calculated by Trotet et al. (2006) from the mineral assemblages and compositions observed in aluminous metapelites using a multi-equilibrium calculation approach (see also [Vidal and Parra, 2000], [Trotet et al., 2001b], [Vidal et al., 2001], [Bosse et al., 2002] and [Parra et al., 2002a]). The calculation involves assemblages made of either chlorite–chloritoid–quartz ± phengite ± sudoite or Chl–Car–Qtz ± sud ± phg ± Cld in the presence of excess water. In the same rock sample or thin section, different *P*–*T* points were obtained by using different generations of phyllosilicates of contrasting compositions that formed at different *P*–*T* conditions in different structural positions. A plot of all *P*–*T* conditions for the same rock samples defines a trend that depicts the local *P*–*T* path. Different *P*–*T* paths, which also reflects the detailed variability of the mineralogical content of the PQ Nappe, were obtained for the four units recognised above that show a contrasting evolution during the retrograde path from peak *P*–*T* conditions of around 18 kbar and 550 °C. Units that have best preserved HP–LT parageneses show a continuous cooling during decompression whereas the Alagonia and Lada units show first an isothermal decompression while the last part of the retrograde path proceeds along a rather warm gradient.

An additional sample taken from the Kithira metapelites of the PQ Nappe shows a *P*–*T* evolution quite similar to the warmest sample of the southern Peloponnese (Fig. 6).

In the southern Peloponnese early studies provided K–Ar ages around 23 Ma for the peak of pressure (Seidel et al., 1982), a finding later confirmed by $^{40}Ar/^{39}Ar$ dating of the phengites of a metagranite included in the Phyllite–Quartzite Nappe of Kithira (approx. 19 Ma, Seidel et al., 2006). Our study using $^{40}Ar/^{39}Ar$ spot fusion ages on micas of the Taygetos and Parnon ranges (Fig. 7 and Fig. 8) shows a distribution of ages mostly between 26 and 13 Ma PQ Nappe.

P–T–t paths can be proposed for each unit (Fig. 6). $^{40}Ar/^{39}Ar$ spot fusion ages recorded by phengites range from 26 to 13 Ma in the Taygetos massif and 50 and 16 Ma for the Parnon massif. The oldest ages, not shown on Fig. 7, are restricted to the easternmost sampling sites in the Neapolis metapelites and the greenschist unit (Alagonia unit). Fig. 7 illustrates the intra- and inter-sample age variations observed in the Taygetos and Parnon massifs. Given the variability of retrograde *P–T* paths displayed on Fig. 6, the coexistence of several mica generations in most samples and the assumed closure temperature for argon in phengite (i.e. most likely in the range 400–450 °C, e.g. Agard et al. (2002) and references therein; Harrison et al., 2009), the age scattering observed within each sample can be interpreted to reflect a combination of cooling and (re)crystallisation effects during rock exhumation from about 450 to 300 °C. For most samples, the age scattering does not exceed 5 Ma and it can be noticed that samples from the same metamorphic zone do not yield results that necessarily overlap. This indicates that the argon behaviour in the studied samples is not only dependent on their cooling history but also on chemical and physical parameters that can potentially act on this behaviour such as local deformation effects or metamorphic reactions, bulk rock composition and permeability, or scale of fluid infiltration ([Scaillet et al., 1992], [Di Vincenzo et al., 2001], [Agard et al., 2002], [Maurel et al., 2003] and [Bröcker et al., 2004]). Therefore, the age distribution can be used only to get a general picture of the thermal evolution of the nappe pile, with a series of internal age variations related to local chemical and physical effects.

In the Taygetos massif the most recent ages are found in the deepest, Meta conglomerate Unit (with car + cld), located immediately above the PK Nappe (sample Tayg.977-5). The data suggest that it was exhumed between 19 and 13 Ma from about 35 to 15 km depth and was still at high pressure while the overlying units with a warmer evolution were already largely exhumed. Given the present-day position of the Metaconglomerate Unit at the base of the PQ stack one can suggest that its continuous cooling during exhumation was due to underthrusting of the cold PK Nappe as has been suggested for the Western Alps (Davy and Gillet, 1986). At the top of the PQ sequence, samples from the garnet-bearing unit (Tayg.9920-1 and Tayg.9713-3) display *in-situ* ages that are significantly older than those from the lower Meta conglomerate Unit, with ages up to 26 Ma. Given the maximum temperature reached by these samples, it is likely that the oldest ages are cooling ages. The maximum pressure was thus reached before ~ 26 Ma. In contrast to the underlying carpholite-bearing and glaucophane or chloritoid-bearing units, the more isothermal exhumation *P–T* path recorded in the garnet unit suggests that cooling below 450–500 °C occurred when the rocks had already been largely exhumed. Cooling ages in the range 26–18 Ma mostly record the low-pressure evolution of the samples. In the intermediate glaucophane or chloritoid-bearing metapelites, sample Tayg.9710-9 yields concordant ages close to 19 Ma that, according to the corresponding *P–T* path, probably mark the end of isothermal decompression from 12 to 5 kbar and the beginning of cooling at 400 °C. Therefore, the Taygetos massif shows a complex imbricate of metamorphic units with warmer, older ones on top, and colder, younger ones at the base, in contact with the Plattenkalk unit.

In the Parnon massif, six samples show $^{40}Ar/^{39}Ar$ ages that are spreading from 50 to 16 Ma, with a complex distribution across the nappe sequence. Sample Par 9926-2 from the intermediate glaucophane or chloritoid unit yields the less scattered ages ranging from 21 to

26 Ma comparable to those obtained in the garnet unit of the Taygetos massif. In the underlying garnet unit of Neapolis, samples NEa 962-H, NEa996 and NEa 962-1 have ages ranging from 21 to 50 Ma which suggest a longer and probably polyphased history of crystallization and cooling, at variance with the chronological records in the Taygetos massif. On top of the massif, two samples from the greenschist unit (Par 9924 and Par 9926-3) yield ages in the range 16–42 Ma that can reflect a polyphased evolution of these metapelites, with the younger ages recording cooling at about 300–350 °C and 10 km depth.

The oldest ages found near Neapolis in greenschist units and garnet-bearing units should now be discussed. Such ages have not been found in Crete and are thus quite unique in the PQ Nappe. The Neapolis metapelites show Eocene–Oligocene ages ranging from 29 to 50 Ma. They are found in the eastern part of the PQ Nappe (or the northern part after restoration), close to the Aegean Sea. Their *P*–*T* evolution follows a path that is not strikingly different from other units but their ages are much older. This behaviour is more typical of the Cycladic Blueschists (e.g. Bröcker et al., 2004 and references therein) than of the PQ Nappe. Their position below the GT Nappe is however not easily compatible with the Cycladic Blueschists that overlie the GT Nappe everywhere else ([Bonneau, 1984] and [Jolivet et al., 2004]). Seidel et al. (1982) already reported 34–59 Ma K–Ar ages for the southern Peloponnese suggesting that this is not a unique observation. These older Eocene ages probably derive from inherited micas that partially preserved their isotopic composition during greenschist metamorphism which is a common situation in high-pressure belts (e.g. [Agard et al., 2002] and [Augier et al., 2005]). We thus consider that these old ages do not reflect the HP stage.

3.3. Zaroukla–Feneos window

In the northern Peloponnese (Fig. 9) the pressure difference across the GT/PQ Nappe contact also exists although it is less severe. The geology of the area has been described in different ways.

The basal contact of the Gavrovo limestones and dolomites is sometimes wrongly presented as the top of the PQ Nappe that would then include the Tyros beds, although they normally belong to the base of the GT Nappe. The detachment mapped by Sorel (2000) in the Zaroukla–Feneos (ZF) window is located between the Tyros beds and the overlying GT dolomitic limestone. When observed, the contact shows a thick zone of cataclastic carbonates. One of the outcrops, along the Kratis valleys shows evidence of N–S stretching and shallow-dipping minor faults within the breccia (Fig. 9). This direction of extension and the presence of shallow-dipping faults are well in line with the interpretation of this contact as a recent detachment ([Sorel, 2000], [Flotté and Sorel, 2001] and [Flotté et al., 2005]).

The main pressure gap encountered in the field is, however, located deeper than this contact. It corresponds to the base of the Tyros beds that rest upon the PQ *stricto sensu* (Fig. 9). Xypolias and Doutsos (2000) have interpreted this contact as a shear zone linked with exhumation, a conclusion that is confirmed by our own observations. We have mapped the contact again and we show a slightly different contour especially in the east near Kastania (Fig. 9). The Tyros beds are represented mainly by low-grade metapelites and poorly metamorphosed volcanic rocks whereas the PQ Nappe is made of more intensely deformed and metamorphosed metapelites and metaquartzites. The contact between the PQ Nappe and the Tyros beds is similar to the Cretan detachment. The upper detachment (Zaroukla detachment) between the Tyros beds and the GT Nappe is more recent and acted only in the brittle crust.

The foliation in the PQ Nappe shows a shallow-dip and is axial plane of flattened folds. A conspicuous stretching lineation, striking NE–SW, is observed throughout the PQ Nappe in the Zaroukla–Feneos window (Fig. 9). Xypolias and Doutsos (2000) attempted a quantification of strain across in the PQ Nappe across the window and they concluded to a gradient of finite strain toward the contact with the Tyros Beds. We follow on their work by an observation of macroscopic kinematic indicators. We found mainly top-to-the NE shear sense but some clear opposite senses have been observed locally mainly along the southern part of the window. Even though a main top-to-the-NE shear sense is probable, a significant component of a more symmetrical regime is present, suggesting that the ZF window is a more symmetrical extensional dome than is the case in Crete. We reach here a conclusion roughly similar to that of Xypolias and Doutsos (2000) although we see a larger predominance of top-to-the NE kinematic indicators and thus a less symmetrical structure.

We have studied the petrology of P–Q metapelites in the Zaroukla–Feneos window with the same method and thermodynamic database as for the southern Peloponnese, estimating the *P* and *T* conditions based upon the equilibrium between chlorite and phengite following the multi-equilibrium method. Despite the scarcity of classical index minerals in the PQ Nappe of the Zaroukla–Feneos window the use of the multi-equilibrium method allows to quantify the *P–0* conditions in a quite robust way. *P–T* estimates in the ZF Window (Fig. 6) suggest maximum *P–T* conditions at around 9–10 kbar and 500 °C in the easternmost outcrops (one sample near Kastania, close to lake Stimfalia) and 5 kbar–550 °C in the west. An additional constraint is given by the analysis of the crystallographic organisation of organic matter with Raman spectroscopy (RSCM Method) following the method developed by [Beyssac et al., 2002] and [Beyssac et al., 2004]. This method gives an estimate of the maximum temperature reached by the sample. Fig. 6 shows a good consistency with the multi-equilibrium method on sample KO 0203. The *P–T* path obtained for the sample near Kastania is similar to that of the warmest units in the southern Peloponnese whereas a sample to the west suggests a still warmer evolution. The result of the RSCM method in itself shows a very different *P–T* evolution in the Zaroukla–Feneos window than in the southern Peloponnese or in Crete. The absence of Fe–Mg carpholite and of any evidence of retrograde one suggests a rather low pressure near Zaroukla despite a temperature as high as 550 °C. The *P–T* gradient thus appears higher there than further south and east.

4. Discussion

Most of the exhumation probably occurred within the subduction channel ([Doutsos et al., 2000] and [Xypolias and Doutsos, 2000]) and before true crustal thinning reached the Peloponnese where the magnitude of post-orogenic extension is quite small. The spatial evolution of *P–T* conditions that appears within the PQ Nappe (Fig. 10) is thus significant of variations in the dynamics of the orogenic wedge or subduction channel. From Crete to the north Peloponnese the *P–T* regime changes gradually from a colder regime in eastern Crete to a gradually warmer regime toward western Crete and toward the north in the Peloponnese. The maximum pressure is similar in Crete and the southern Peloponnese, but the maximum temperature is offset by ~ 100 °C using similar estimation methods. The Zaroukla–Feneos window completes this pattern with a rather high *T*/*P* ratio. From east to west the temperature at peak of pressure increases and the retrograde path gets progressively warmer.

In parallel with this thermal evolution the kinematic pattern also changes. In Crete kinematic indicators along the detachment are simple and mostly top-to-the-north ([Fassoulas et al., 1994], [Jolivet et al., 1994b] and [Jolivet et al., 1996]), and sedimentary basins in the Miocene developed concurrently with the activity of the detachment ([Ring et al., 2001a], [van Hinsbergen and Meulenkamp, 2006] and [Seidel et al., 2007]). In the southern Peloponnese the pattern of exhumation and the kinematics of detachments seem more symmetrical although large pressure gaps are still observed. In the northern Peloponnese extension and exhumation seem to be less intense with a smaller pressure gap and quite symmetrical kinematics.

An additional important observation is the absence of HP parageneses in the upper part of the PK of the southern Peloponnese showing that this unit has been buried much less deep (not more than 20–25 km for the base of the PK Nappe, Blumör et al., 1994) there than in Crete suggesting that the overall shortening of the Apulian platform was larger in the centre of the Hellenic arc than in the Peloponnese.

Among the main parameters that control the thermal regime of the subduction channel the velocity of subduction and exhumation is predominant. Exhumation of HP–LT rocks in these regions occurred while the Aegean domain was actively extending in the backarc domain, faster along the Crete–Cyclades transect than further west. Depending upon the efficiency of slab retreat and backarc extension, the subduction channel can be more or less constrained ([Beaumont et al., 1999] and [Jolivet et al., 2003]) and the internal circulation of material into it more or less easy. North of Crete, backarc extension was more efficient than in the internal zones of the Peloponnese and the exhumation of the PQ Nappe appears facilitated by an active detachment at the top that removed the overburden quite efficiently, which could possibly explain that the subduction channel was cold, whereas in the Peloponnese exhumation was more difficult because backarc extension was less active and prevented a fast exhumation in a more tightly constrained subduction channel. Exhumation was even more difficult in the northern Peloponnese and the overall *P*–*T* regime thus appears still warmer.

This potential effect is reinforced by a more efficient subduction south of Crete, the addition of the Africa–Eurasia convergence and of slab retreat (or backarc extension) leads to a faster rate of African lithosphere consumption in the mantle (Jolivet et al., 2008). A more efficient subduction can explain the deeper conditions recorded in the PK Nappe in Crete than in the Peloponnese. A higher subduction rate will furthermore lead to a colder slab below the subduction channel in Crete than in the southern Peloponnese (Fig. 11).

To what extent later extension played a role in the final exhumation as in the Cyclades is difficult to assess and further work is needed to answer this question. This conclusion strongly limits anyway the amount of a possible relative normal sense displacement along the detachment in the Pliocene and Quaternary (Papanikolaou and Royden, 2007), most of the displacement was achieved during exhumation and not during whole crust extension. This is a situation different from the Cyclades where a strong post-orogenic crustal thinning has reactivated syn-orogenic detachments.

5. Conclusion

We show in this paper that the *P*–*T* conditions and the internal kinematics of an accretionary wedge or subduction channel can change significantly along strike depending upon the kinematic boundary conditions. Using the example of a single tectonic unit with a monotonous lithology, the Phyllite–Quartzite Nappe, we document an evolution toward lower temperatures when the rate of slab retreat increases. This can be due to an easier circulation of the subducted material within a less constrained subduction channel and to a higher rate of subduction of a cold retreating slab. An overall difference of ~ 100 °C is recorded between the northern Peloponnese and Crete for the peak of pressure (~ 16–18 kbar) and the shape of *P–T* paths is also affected. These observations can now be used as proxies for numerical models.

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Figures

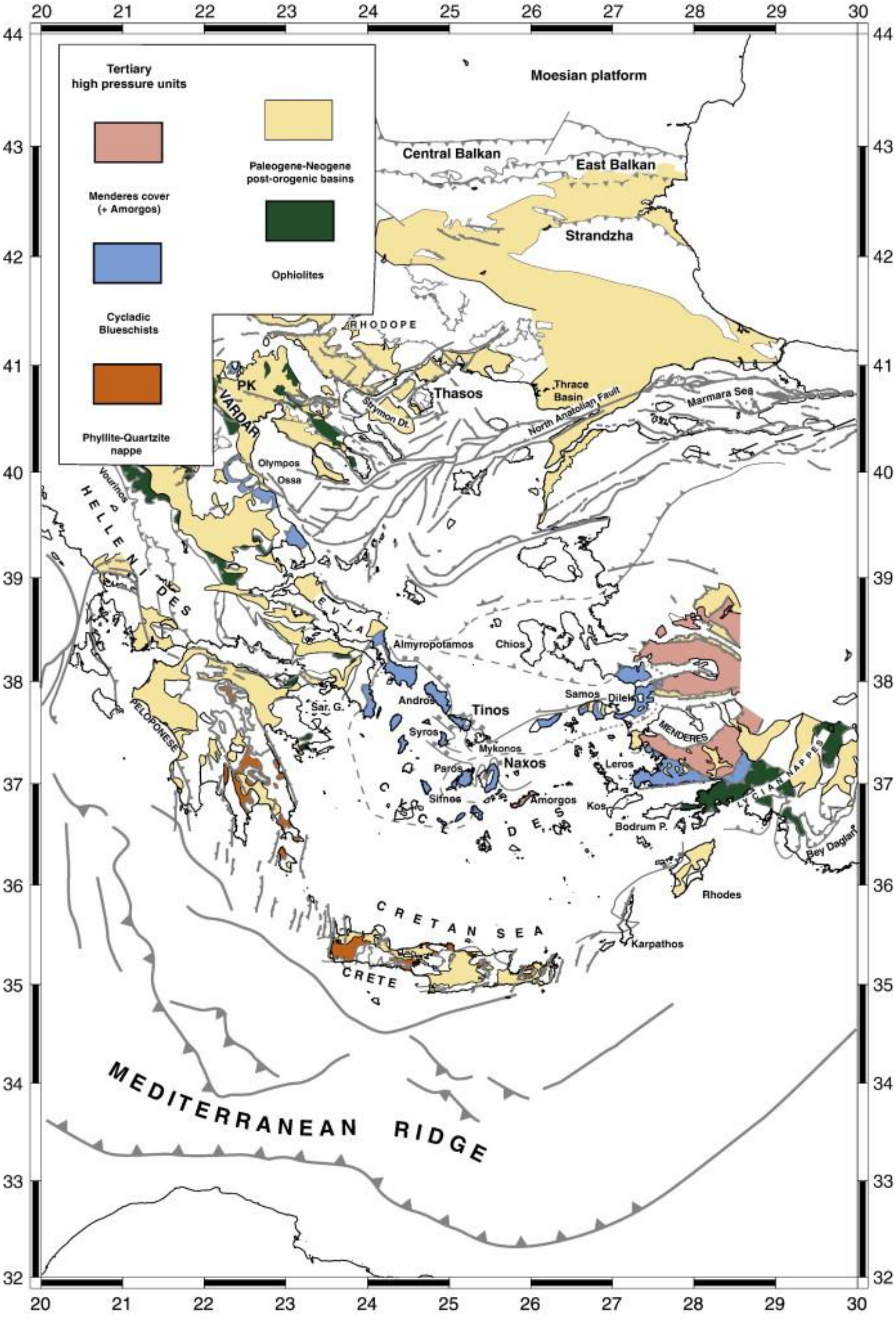

Fig. 1. Tectonic map of the Aegean region showing the main faults, and the main HP–LT metamorphic units of Cenozoic age, the Cycladic Blueschists and the Phyllite–Quartzite Nappe.

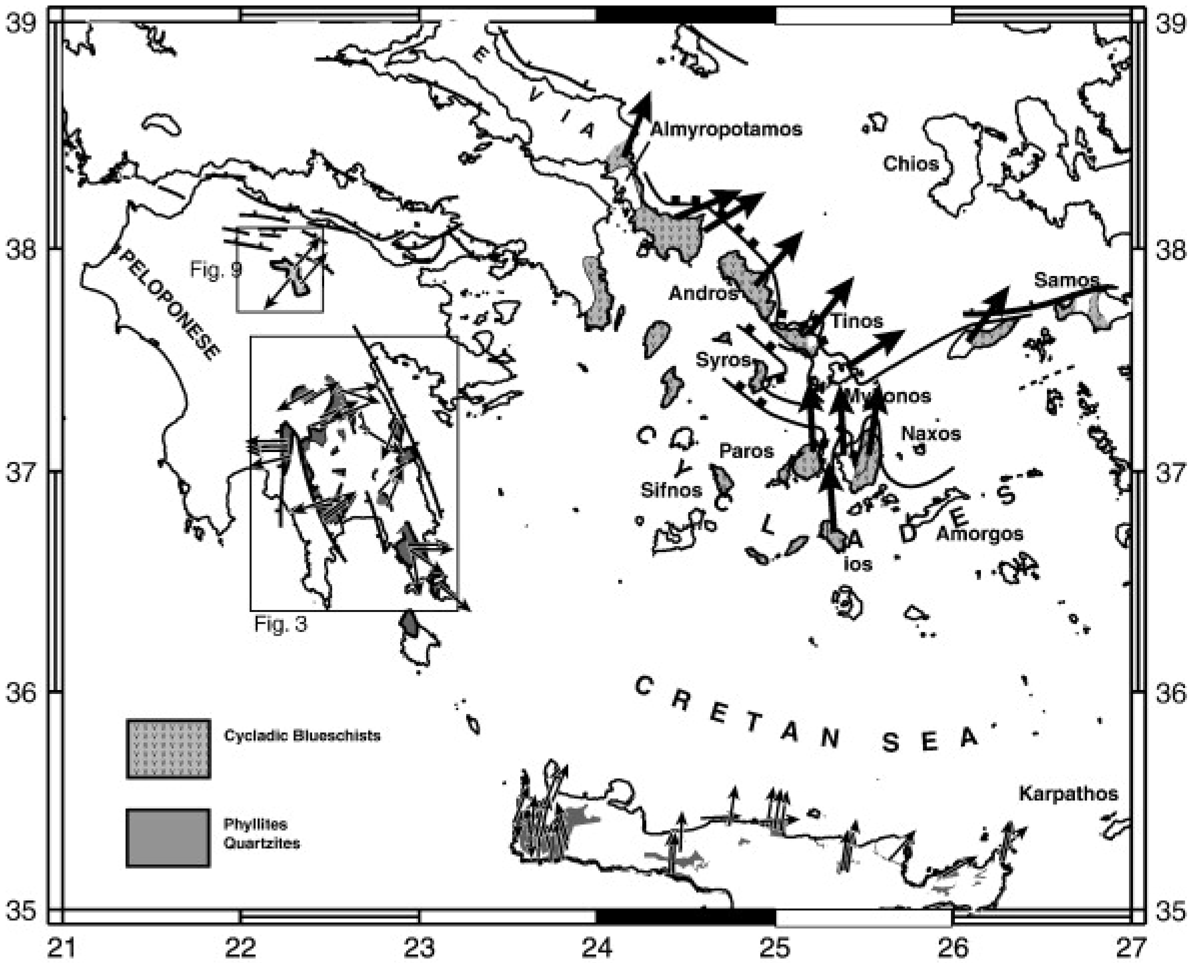

Fig. 2. Map of the Cyclades, Peloponnese and Crete showing the main outcrops of the HP–LT metamorphic rocks. Arrows shows the direction of retrograde stretching lineations of Oligo-Miocene age and the associated kinematic indicators after ([Gautier et al., 1993], [Jolivet et al., 1994a], [Jolivet et al., 1996], [Doutsos et al., 2000], [Trotet, 2000] and [Xypolias and Doutsos, 2000]). Large arrows for the Cyclades, smaller arrows for the PQ Nappe.

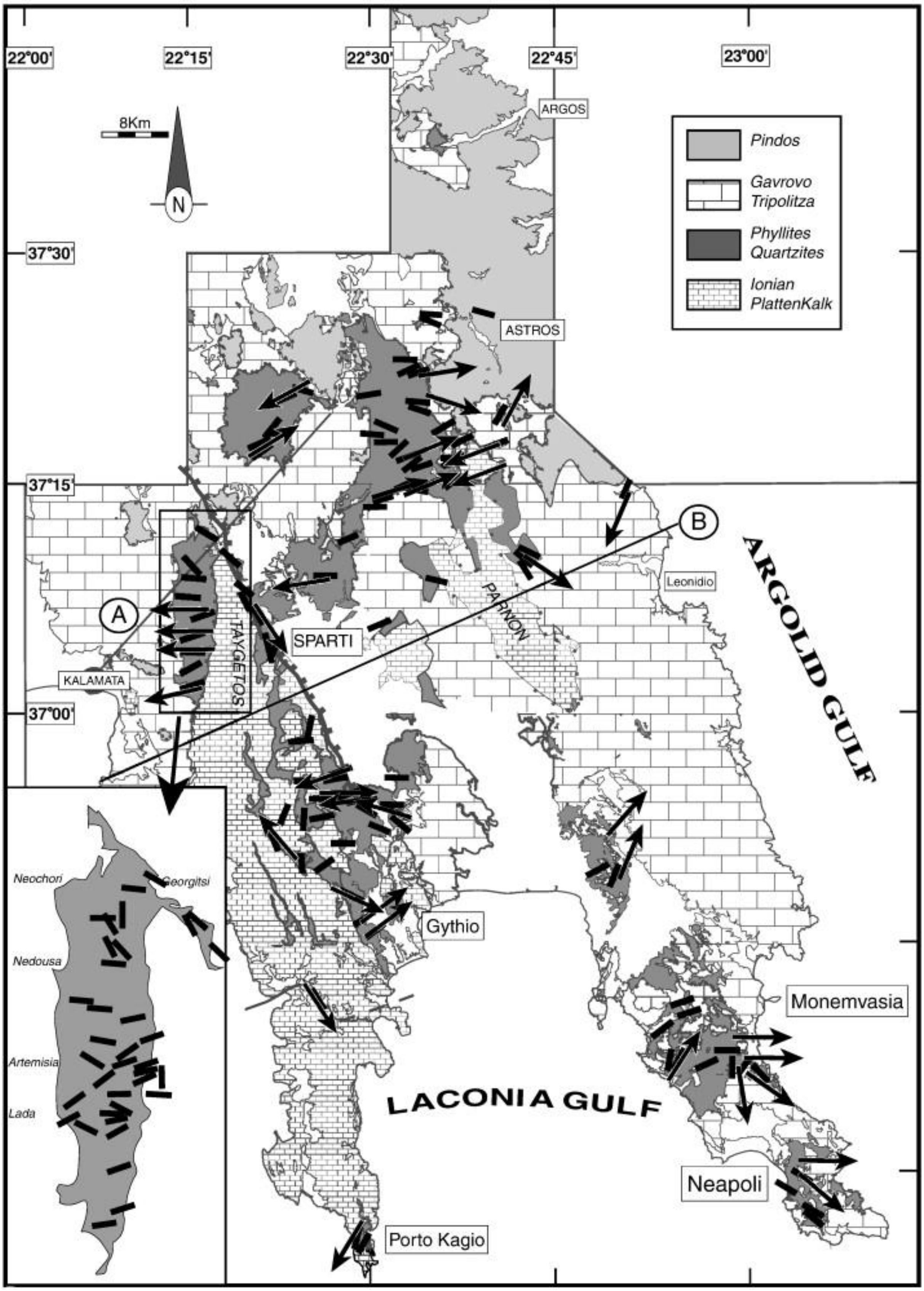

Fig. 3. Tectonic map of the southern Peloponnese showing the position of the Phyllites–Quartzites Nappe and Oligo-Miocene retrograde kinematic indicators. Black bars represent the strike of stretching lineation and black arrows shows the associated shear of shear when known.

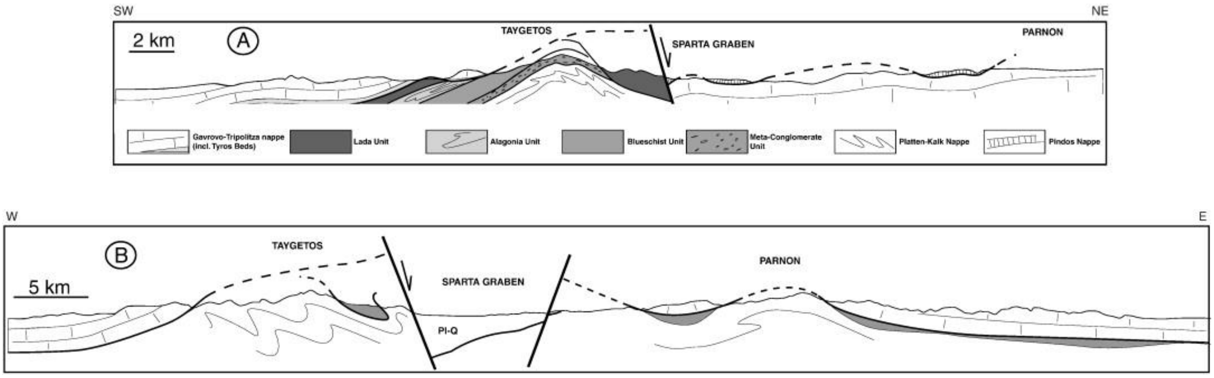

Fig. 4. Cross-sections (see location on Fig. 3) of the Taygetos and Parnon ranges showing the relations between the various units of the Phyllite–Quartzite Nappe.

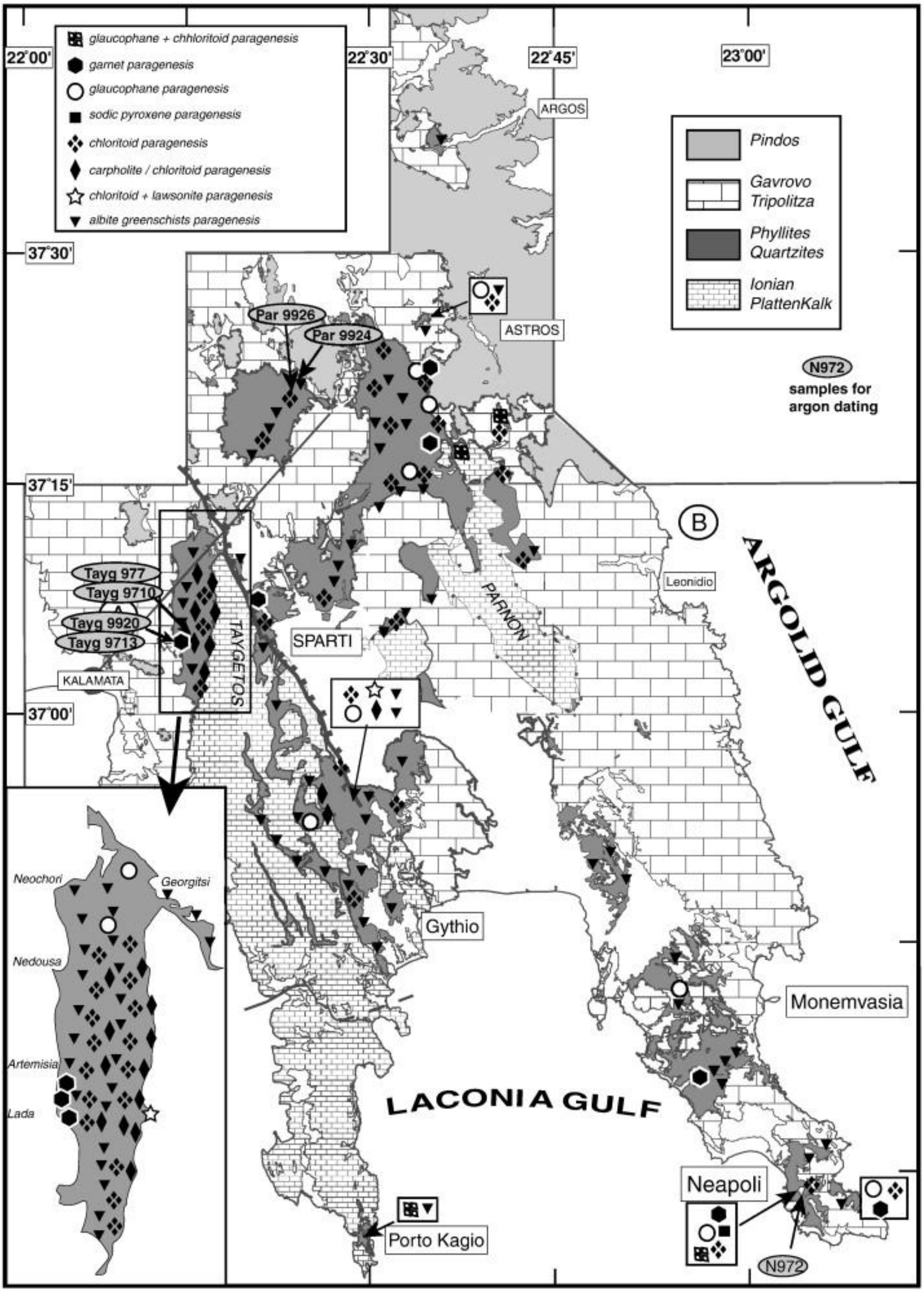

Fig. 5. Tectonic map of the southern Peloponnese showing the distribution of parageneses within the PQ Nappe and the location of samples used for spot fusion $^{40}Ar/^{39}Ar$ dating. (A) and (B) indicate the cross-section lines of Fig. 8.

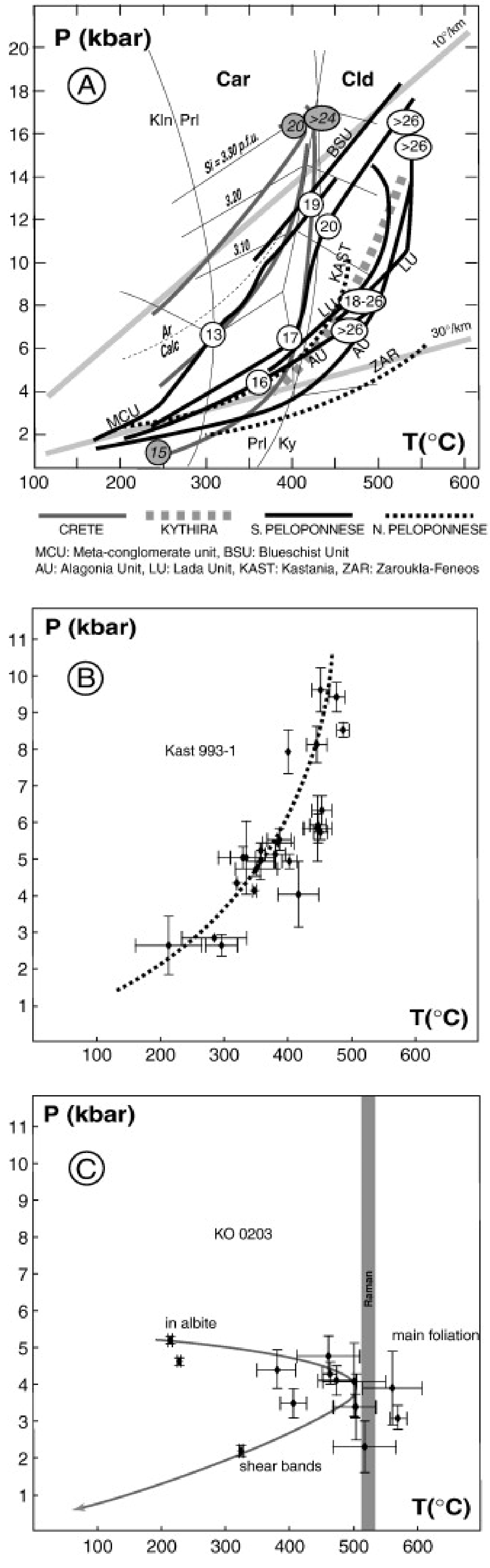

Fig. 6. (A) *P–T* paths for the Phyllite–Quartzite nappe in Crete, Kithira and Peloponnese after [Jolivet et al., 1996], [Trotet, 2000] and [Trotet et al., 2006]) and this work projected on the FMASH *P–T* grid

of Vidal et al. (1992) for XMg-carpholite = 0.5 and XMg-chloritoid = 0.2. The Si-isopleths of phengite calculated by Goffé and Oberhänsli (1992) are also reported in the carpholite and chloritoid stability fields. Numbers in circles give the $^{40}Ar/^{39}Ar$ ages (see text for further explanation) (step-heating for Crete — grey circles — after Jolivet et al. (1996), and spot-fusion ages for the Peloponnese (this study — white circles). Four units are recognized, (1) blueschist-facies metaconglomerates and micaschists with Fe–Mg carpholite and chloritoid hereafter named the Metaconglomerate Unit (MU), (2) blueschist-facies micaschists with chloritoid and lenses of glaucophanites, hereafter named the Blueschist Unit (BU), (3) greenschist-facies micaschists with chloritoid and albite (Alagonia unit, AU). In the Taygetos one more unit comes near the top of the stack below the GT Nappe, it contains (4) HT blueschist-facies micaschists with glaucophane and garnet (Lada unit, LU). (B) *P–T* conditions recorded in sample Kast993-1 near Kastania in the eastern part of the Zaroukla–Feneos window. (C) *P–T* conditions recorded in the Zaroukla–Feneos window for the Phyllite–Quartzite (grey boxes, sample KO 0203, western part of the window); the same methodology has been used. Maximum temperature conditions estimated by the Raman spectra of carbonaceous material confirm a maximum temperature of ca. 500 °C. See samples location on Fig. 3 and Fig. 9.

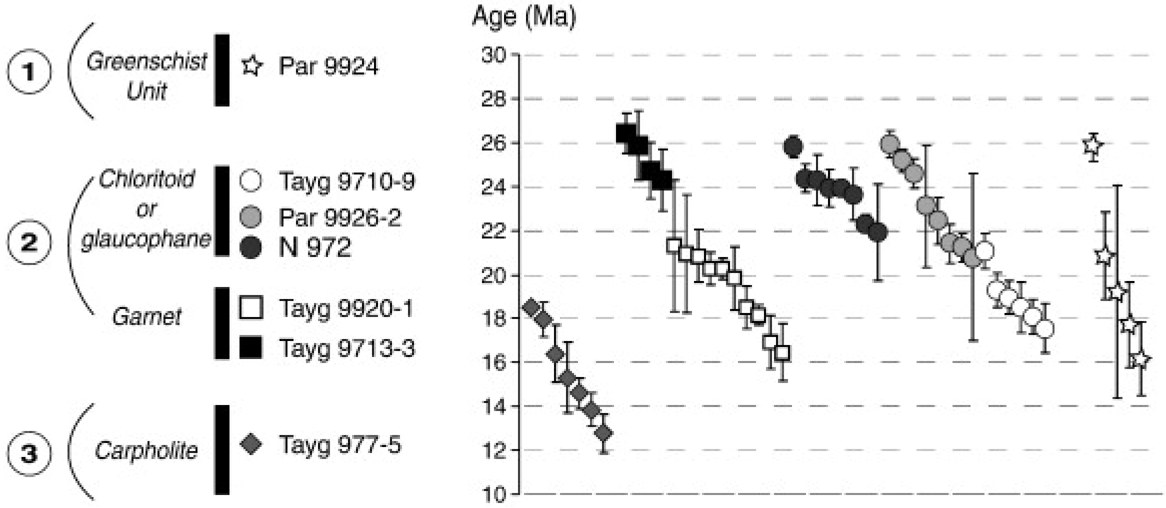

Fig. 7. Spot fusion Ar/Ar ages in the Phyllites–Quartzites nappe of the Southern Peloponnese. Samples Par 9924 and Par 9926-2 are from the Parnon massif, Tayg 9710-9, Tayg 9920-1, Tayg 9713-3 and Tayg 977-5 from the Taygetos massif and N 972 (Glaucophane schist) from Neapolis. See sample location on Fig. 5. During this work, rock sections of 1 mm thick were used for in-situ laser probe $^{40}Ar/^{39}Ar$ dating because of the complex metamorphic evolution of the PQ Nappe and the general occurrence of more than one mica population in the studied samples. We follow a method previously used by [di Vincenzo et al., 2004], [Agard et al., 2002], [Mulch, and Cosca, 2004] and [Augier et al., 2005]. The analytical procedure follows that previously detailed in Agard et al. (2002; Augier et al., 2005). Samples of about 1 cm square and 1 mm thick were selected after careful examination of their mineralogy in thin sections made on the same rock slab. These samples were irradiated in the McMaster nuclear reactor (Canada) together with several aliquots of the hornblende standard MMHb-1 (520.4 +/− 1.7 Ma; Samson and Alexander, 1987). Argon has been released from the irradiated samples using an argon laser operating in a semi-pulsed mode with principal wavelengths at 488 and 514 nm. In order to get a sufficient amount of argon for mass spectrometry analysis, a volume of about 200 × 200 × 20 μm was molten for each age determination which can correspond to the degassing of one or several mica grains depending on their size. After gas cleaning, the argon isotopic composition was measured on a MAP 215-50 mass spectrometer by peak jumping from mass 40 to 36. Ages are reported with one sigma uncertainty and were obtained after correction with blanks, mass discrimination, decay effects and nuclear isotopic interferences. Data showing a large contribution of Ca-derived ^{37}Ar and Cl-derived ^{38}Ar due to contamination by adjacent phases or fluids have been excluded from these results.

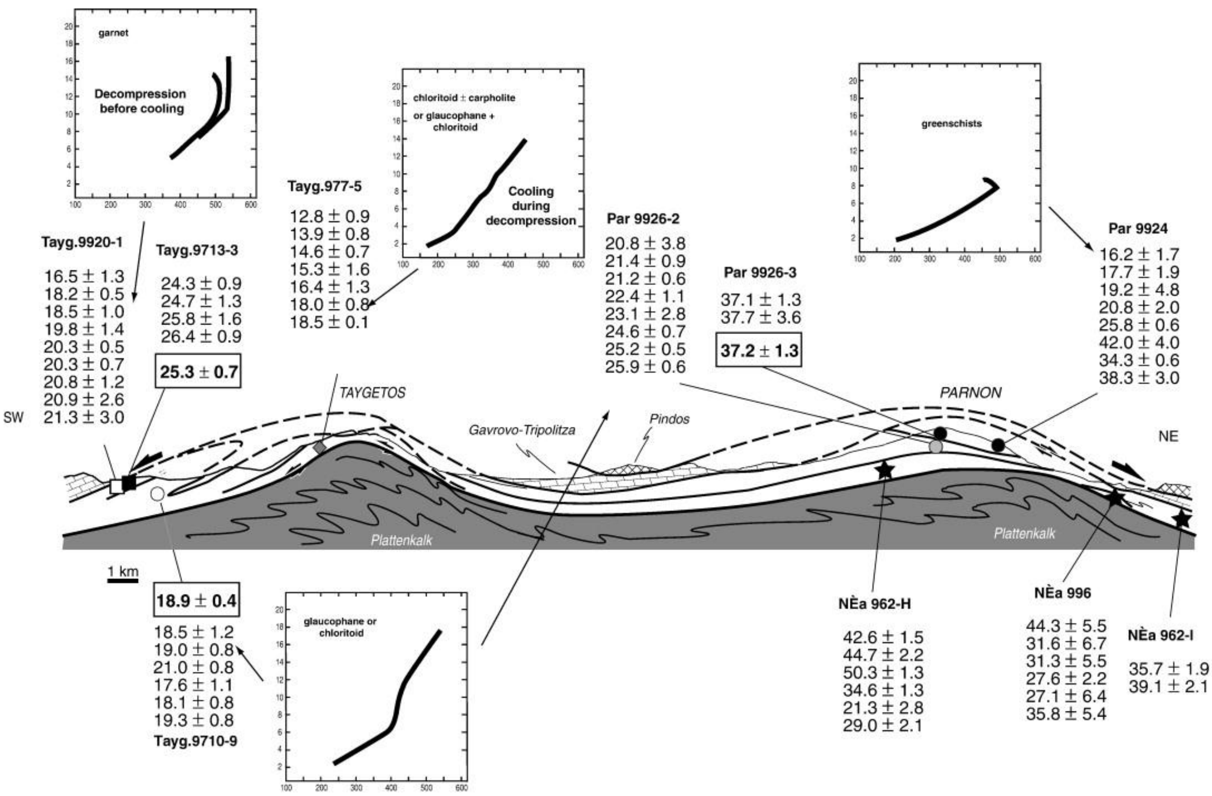

Fig. 8. Schematic cross-section (see location on Fig. 3) of the Taygetos and Parnon ranges showing the *P–T* evolution of various units as well as the spot fusion Ar/Ar ages and some average values (boxes). Dots, squares and stars refer to the samples plotted on Fig. 7. For clarity, recent normal faults have been omitted.

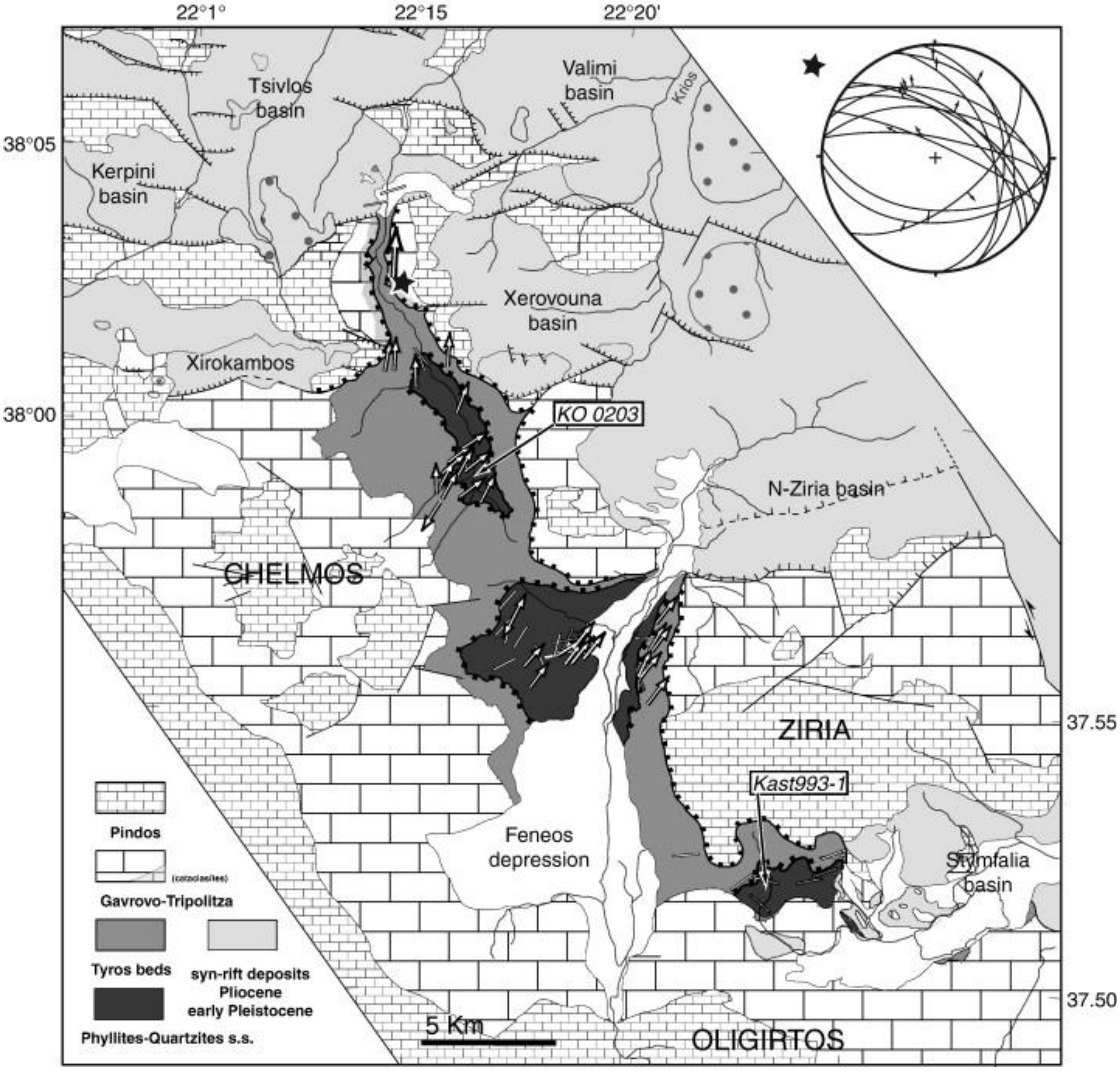

Fig. 9. Geological map of the Zaroukla–Feneos window in the northern Peloponnese (see location on Fig. 2). We distinguish the true metamorphic Phyllite–Quartzite Nappe (lowermost unit) from the Tyros beds that make the lower part of the Gavrovo–Tripolitza Nappe (see also Flotté et al.; Xypolias and Doutsos, 2000). The direction of retrograde stretching lineations and shear sense when available is also shown. Small arrows represent the shear direction in the PQ Nappe and the large arrow the transport direction on the Zaroukla detachment (next to black star). Inset, Stereographic projection plot of normal faults and their striation in the basal breccia of the GT Nappe showing an overall N–S direction of stretching. KO 0203 and Kast993-1 are the samples used for *P*–*T* estimates (Fig. 6).

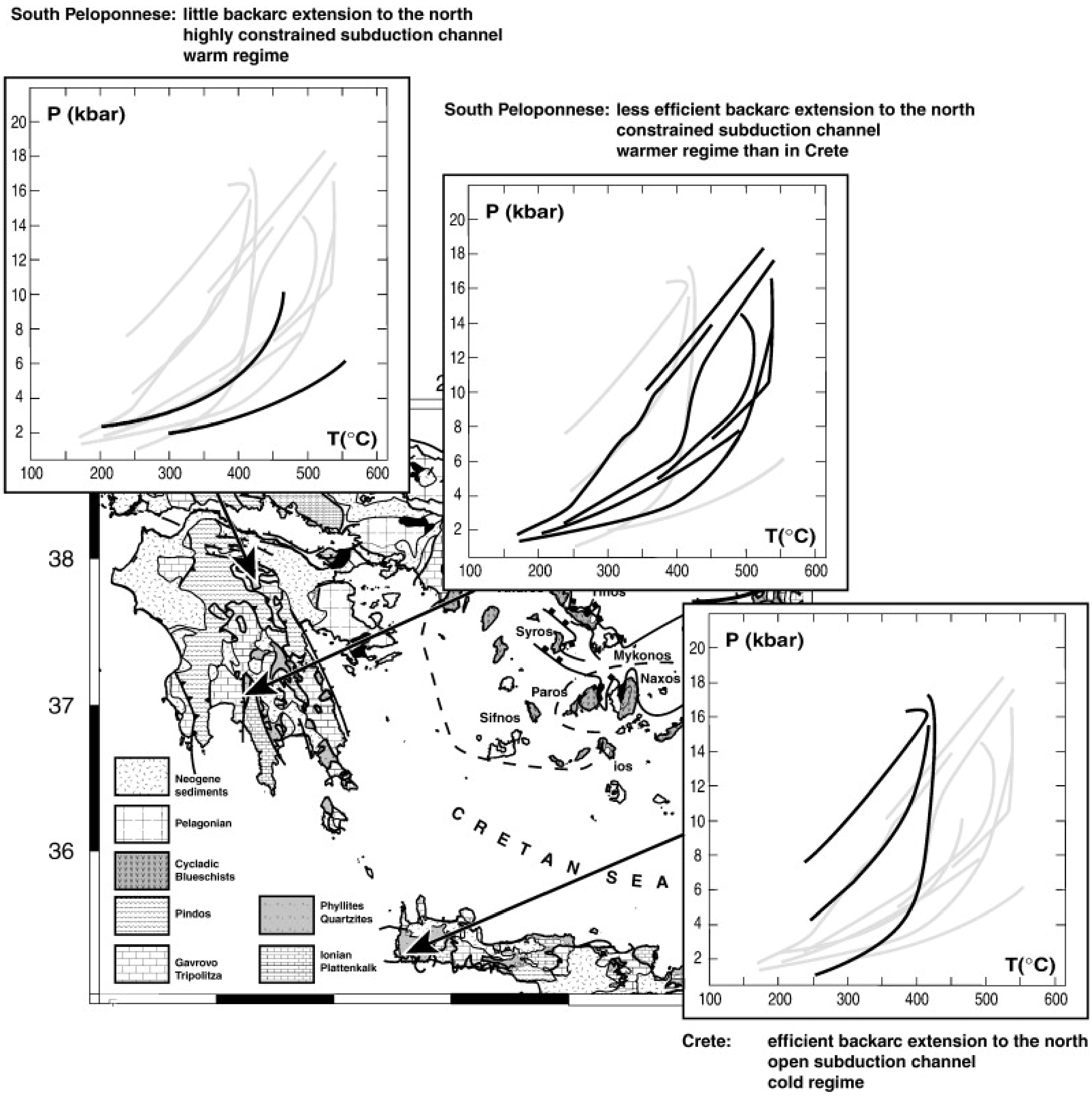

Fig. 10. Synthesis of the *P–T* evolution of the PQ Nappe from Crete to the Peloponnese. *P–T* conditions are cold in Crete and evolve toward warmer conditions westward and northward. Kinematic boundary conditions also change with a more efficient backarc extension north of Crete than east of the Peloponnese. This change in *P–T* conditions is tentatively attributed to a more open subduction channel in Crete and a progressively more constrained one toward the northern Peloponnese because a more active backarc extension in the centre of the Aegean Sea.

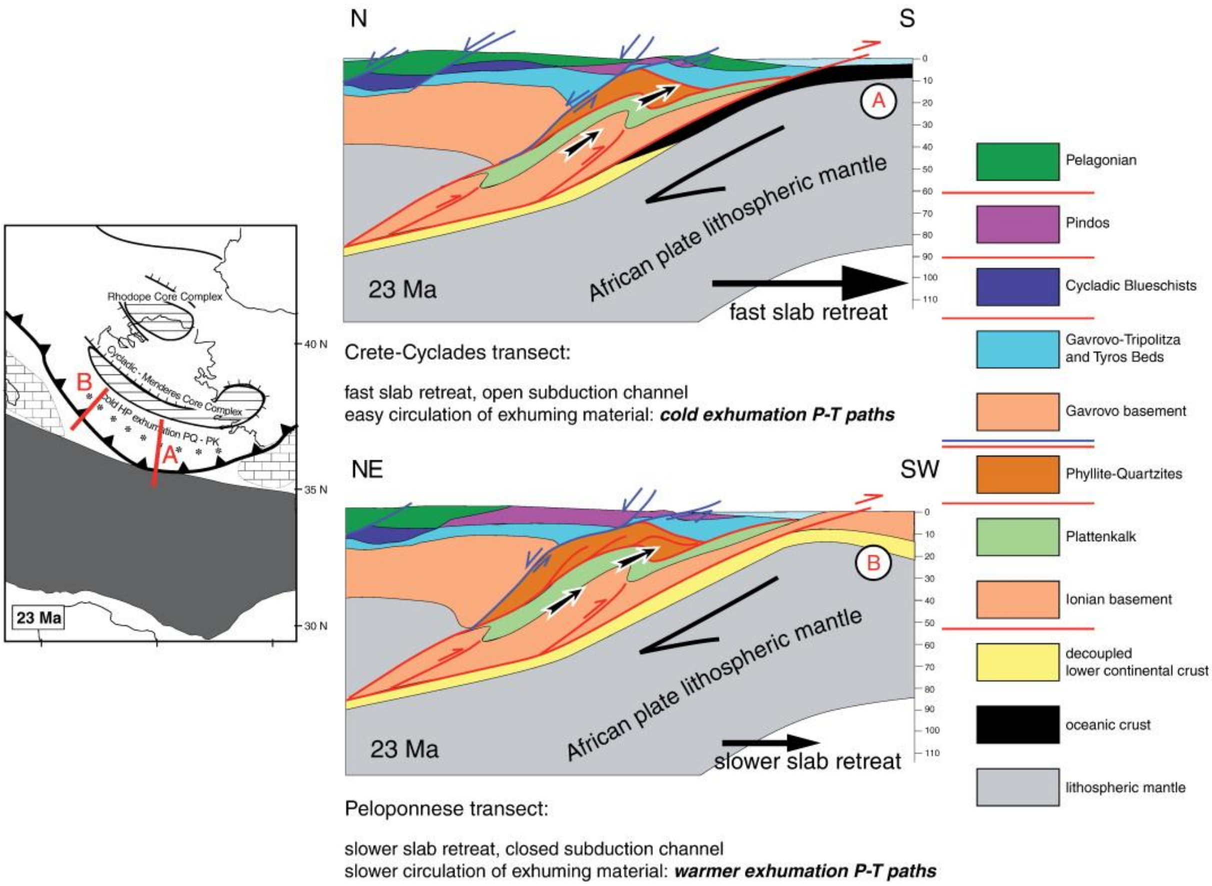

Fig. 11. two schematic cross-section of the Hellenic subduction zone at 23 Ma, one across the Peloponnese (B) and one through Crete and the southern Cyclades (A) putting the along strike evolution of *P*–*T* conditions into its geodynamic framework. (A) is a detail of the 23 Ma stage in the reconstructions proposed by Jolivet and Brun (2008). At this longitude, in the center of the Hellenic arc, slab retreat reaches its maximum velocity leading to colder conditions; furthermore, the subduction channel is open, letting the subducting material circulate easily leading to a cold subduction channel and accretionary complex. Panel (B) is located farther west where slab retreat is less efficient and the rate of subduction thus smaller. Continental material is still present at the trench. Because slab retreat is slower the system is colder; the subduction channel is also more constrained and the material inside circulates less easily leading to warmer conditions. The subduction channel and the accretionary complex are thicker.